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1     **Eskers in a complete, wet-based glacial system in the Phlegra Montes region, Mars**

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9

10    **Abstract**

11    Although glacial landsystems produced under warm/wet based conditions are very common on  
12    Earth, even here, observations of subglacial landforms such as eskers emerging from extant  
13    glaciers are rare. This paper describes a system of sinuous ridges emerging from the *in situ* but  
14    now degraded piedmont terminus of a Late Amazonian-aged (~150 Ma) glacier-like form in the  
15    southern Phlegra Montes region of Mars. We believe this to be the first identification of  
16    martian eskers that can be directly linked to their parent glacier. Together with their contextual  
17    landform assemblage, the eskers are indicative of significant glacial meltwater production and  
18    subglacial routing. However, although the eskers are evidence of a wet-based regime, the  
19    confinement of the glacial system to a well-defined, regionally significant graben, and the  
20    absence of eskers elsewhere in the region, is interpreted as evidence of sub-glacial melting as a

21 response to locally enhanced geothermal heat flux rather than climate-induced warming. These  
22 observations offer important new insights to the forcing of glacial dynamic and melting  
23 behaviour on Mars by factors other than climate.

24

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26 **Keywords:** Mars; glacier; eskers; wet base; geothermal control.

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## 1. Introduction

The Phlegra Montes upland extends NNE-SSW over 1000 km, between 30°N and 52°N on Mars (Fig. 1). Southern Phlegra Montes, 560 km east of Hecates Tholus on the Elysium rise, is 90 to 180 km wide, overlooking the Elysium Rise to the west and sloping towards a flanking N-S trending piedmont basin to the east. The upland relief is dominated by rounded peaks, intervening valleys and basins, some partially occupied by icy fills (Safaeinili et al., 2009; Dickson et al., 2010). The valley fills are longitudinally lineated ('lineated valley fills' or LVF; Squyres, 1979), like the basin fills exhibiting ridges, troughs and lobes. These morphologies, suggesting flow over or around obstacles in response to changes in underlying slope, are typical of 'viscous flow features' (VFF; Milliken et al., 2003). These characteristics, together with associated erosional (mainly upland) and depositional (including piedmont) landforms such as moraine-like ridges are regarded as evidence that VFF and LVF are glaciers, or glacier-like forms (GLF; Souness et al., 2012; Hubbard et al., 2014), formerly thicker and more extensive. The crater retention age of these landforms indicates they formed over the past ~600 Ma (Kress et al., 2010; Fassett et al., 2014), most recently in the Late Amazonian (Milliken et al., 2003; Hubbard et al., 2011; Souness et al., 2012; Hubbard et al., 2014). In common with other low plains bounding uplands, the Phlegra Montes piedmont is characterized by lobate debris aprons (LDA), accumulations of ice mantled by lithic debris (Kochel and Peake, 1984; Holt et al., 2008; Parsons et al., 2011; Fastook et al., 2014).

Fig. 1 here.

51 The majority of observational glaciological and landform evidence shows that extant martian  
52 glaciers/GLF are cold/dry based (Hubbard et al., 2011; Hubbard et al., 2014)and (were) dynamic  
53 by virtue of creep (Milliken et al., 2003; Parsons et al., 2011); the landform evidence for relict  
54 glacial process-environments suggests this has been characteristic of Amazonian glaciation.  
55 Observations of landforms in contextually consistent landsystems diagnostic of warm/wet  
56 based glacial regimes on Mars, especially eskers,are rare in comparison to the widespread  
57 presence of glaciers.Kargeland Strom (1991), however,were confident that many sinuous, often  
58 branching, ridges on Mars are eskers that, as on Earth, display a wide size range and ridge-  
59 network variety, ranging from single to branching to arborescent and braided. More recently,  
60 several researchers have concluded that sinuous ridge systems in Dorsa Argentea (Head, 2000)  
61 and Argyre Planitia(Banks et al., 2009; Bernhardt et al., 2013) are eskers, the latter reflecting  
62 subglacial routing of pressurized meltwater, generated both supraglacially and englacially,  
63 through Rothlisberger (R) channels cut upwards into extensive Hesperian glacial ice. On Earth,  
64 landsystems produced under warm/wet based conditions, including organized englacial to  
65 subglacial meltwater routing and sediment flux,are very common, dominating the landscapes of  
66 many deglaciated areas. However, observations of subglacial landforms in a state of emergence  
67 from degrading but extant glacial ice are rare, even on Earth, and previously un-reported on  
68 Mars. This paper describes a system of sinuous ridges emerging from the degraded piedmont  
69 terminus of a LVF/GLFin the southern Phlegra Montes region.Based on analysis of the  
70 landsystem as a whole, the conclusion is that these landforms are eskers emerging from a  
71 decayed glacial margin. Together with theircontextual landform assemblage, the eskers are

indicative of significant glacial meltwater production and subglacial routing – by definition evidence of a wet-based regime. Whether they are indicative of acclimatically-determined warm-based regime is discussed, and alternatives considered.

## **2. Approach**

### **2.1 Data**

Covering the study area, a mosaic of seven, 6 m/pixel, georeferenced ConTeXt camera (CTX, Malin et al., 2007) images (Appendix 1 and 2) was constructed using ArcGIS. Other data included a High Resolution Stereo Camera (HRSC; Neukum and Jaumann, 2004) image and its associated Digital Elevation Model (DEM), and gridded Mars Orbiter Laser Altimeter (MOLA; Zuber et al., 1992) topography data. The MOLA data have low spatial resolution (~500m gridding) but high vertical precision (~1m). The HRSC DEM has higher spatial resolution (75m gridding) but vertical precision similar to the spatial resolution of the original image data (i.e. about 12 m).

### **2.2 Determining the age of the system using impact crater size-frequency statistics**

Planetary surfaces can be dated using impact crater size-frequency distribution data, although this becomes complicated following resurfacing, surface modification or downwasting (Michael and Neukum, 2010). It is also difficult when considering small areas in which insufficiently large populations of craters have accumulated to provide statistically reliable ages (Warner et al., 2015). Both problems apply to the land system in Phlegra, as the LVF has probably downwasted

over time through loss of ice, and the extent of the LVF and associated landforms is only a few 100km<sup>2</sup> in areal extent. To estimate the formation and modification ages of the system, the size-frequency distribution of impact craters was measured for various sub-regions using the ArcGIS add-ons *Cratertools*(Kneissl et al., 2011) and plotted using the tool *CraterStats*(Michael and Neukum, 2010).

The *CraterTools* 3-point method was used to digitise the rims of all visible impact craters. Crater discrimination was complicated by the many rimless circular features on the LVF - most are probably degraded craters but the relationship between their present and original diameter is unclear. We counted all the crater-like circular features, recording their current diameter. Consequently, this crater-count errs towards overestimating the crater retention age of the LVF. However, this is balanced to an unknown degree by the likely loss of many craters from the LVF surface due to sublimation and flow deformation. Hence, owing to the small count area, the low number of craters counted and surface modifications, the count data for the LVF especially give only a first order approximation of the age.

### **3. Observations: landsystem components.**

The main landsystem components (Fig. 2a) are a parallel-sided, trough-like valley, striking WNW-ESE through the far-southern Phlegra Montes upland (Fig.2, Zones 1-2), and a relatively shallow trench extending through the eastern piedmont directly along strike from the upland valley (Zones 3-4). The valley is occupied by a diffluent LVF, descending to the western and eastern piedmonts (Fig.2b, Fig 3a). Tributary troughs containing backwashed LVF, which no

longer reach their confluence, incise the plateau overlooking the valley (Fig. 3b). The LVF is patterned by lineated, viscous features - the surface expression of glacier-like flow that characterises GLFs (Milliken et al. 2003; Souness et al., 2012). East of the LVF, a pitted zone follows the same topographic trend (Fig. 2, Zone 2). The trench incising the eastern piedmont (Fig. 2, Zones 3 and 4) is occupied by a fill that varies from hummocky and intensely pitted to longitudinally-furrowed. The trench is laterally bounded by higher, pitted piedmont surfaces and terminates at a low-lying, distal basin occupied by a fractured, level fill (Fig. 2, Zone 5). A complex of sinuous ridges is located at the terminus of the trench fill. The piedmont surface close to the upland is mantled by pitted mass-wasting deposits shed from the bounding slopes. The distal piedmont is less intensely pitted but, like those of the proximal piedmont, the pits are themselves internally textured by smaller pits. The upland LVF descending eastward, the piedmont trench-fill (PTF) and its terminal zone sinuous ridges are the focus of this paper.

Fig. 2 here

### 3.1 Zone 1

The apex of the LVF descending eastward occupies the widest part of the upland valley: ~11 km across (Fig 2). This zone is a flattened dome with a subtle, hummocky surface. The depth of the LVF is unknown. Although there are SHARAD (Seu et al., 2007) RADAR profiles across the system, showing a possible sub-surface reflector, their usefulness is eliminated by the presence



of off-nadir ‘clutter’. The LVF narrows downvalley to the east, becoming 7.9 km wide at its present terminus, 15.5 km from the apex. The LVF is not a simple form but consists of asymmetric lobes converging from the valley sides to its mid-line (Fig. 3a). The LVF surface is longitudinally lineated, with alignments replicating the changing asymmetry of the constituent lobes as the valley widens or narrows. Degraded ring-mold impact craters, thought to indicate impact into near-surface ice (Kress and Head, 2008), and a few fresh craters occur on the LVF surface (Fig 3a). Zone 1 terminates at a 900 m-long, back-sloping belt of irregular pits and blocky hummocks spanning the valley, here ~7 km wide (Fig 4a). Individual pits are at most 100 – 300 m wide, hummocks 100 – 250 m, and the entire assemblage stands up to ~80 m higher than the LVF terminus.

Fig 3 here.

### 3.2 Zone 2

The first pitted band at the foot of the LVF marks the beginning of a 3.3 km-long reach of valley fill (Fig 4a) terminating at another cross-valley belt (pit band 2) of 100m-scale pits and hummocks. This belt marks the start of a 6.8 km-long reach of pitted, hummocky fill, bounded by two prominent converging headlands (A and B, Fig. 4a) that define a narrow valley (head at ‘P’, Fig. 4a). The headlands were formed by the breaching of an originally continuous transverse bedrock ridge at its lowest point (‘Q’, Fig. 4a). The breach is ~0.8 km wide and ~30 m deep from

the level of its shoulders. Although the lateral margins of the Zone 2 hummocky and pitted fill are generally in contact with both valley sides, a small valley extends from the southern limb of the breached ridge for ~4 km into the LVF. This re-entrant (X to Yin Fig. 4a) is characterised by longitudinally linked alcoves, up to 1.5 km wide, and a sharply defined sinuous valley axis.

Fig. 4 here

### 3.3 Zone 3

Beyond the breached bedrock ridge delimiting Zone 2, for ~30 km along the same strike as the upland valley, the eastern piedmont is indented by a 3 – 5.5 km-wide trench containing the PTF. In the proximal eastern piedmont, the trench is laterally bounded by well-defined walls, rising up to 100 m above the intensely pitted, hummocky PTF (Fig. 4b) and indented by erosional alcoves, best developed on the north-facing wall (white arrows, Fig. 4b). A suture-like crease, expressing the axial continuation of the ridge-breach in Zone 2, continues into the Zone 3 PTF (black arrows, Fig. 4b). Zone 3 has a shallow eastward-dipping slope, compared with the steep, convex slopes that mark the eastern half of Zone 2 (Fig. 2b).

### 3.4 Zone 4.

171 Zone 4 encompasses the PTF traversing the distal eastern piedmont. The transition from Zone 3  
172 is subtle, but distinct, marked by a change in the pattern and albedo of the hummocky pitted  
173 fill. The zone has three sub-zones, each showing increasing amounts of connectivity along  
174 channel-like systems.

175 **Zone 4a:** From the end of Zone 3 for ~4.8 km, the northern and southern PTF margins are  
176 dominated by relatively low albedo, convex-up, longitudinally-furrowed, undulating swaths up  
177 to ~1 km wide (FN and FS, Fig 5a). Most of the central tract along this PTF reach is disrupted into  
178 a curving series of alternating bulbous steps and depressions. MOLA and HRSC DEM data  
179 suggest this tract is ~10m lower than FN and FS. The longitudinally-furrowed swaths, therefore,  
180 are flanking platforms bounding the disrupted tract. Both bounding platforms terminate at a  
181 poorly-defined, low, curving scarp marking the transition to Zone 4b but which, in the central  
182 disrupted tract, cuts 1 km back into Zone 4a. Some longitudinal furrows appear to cross into  
183 Zone 4b, whereas the disrupted steps and depressions terminate at, or form, the disrupted  
184 scarp. Although at the limits of HRSC vertical resolution, the topographic interpretations are  
185 supported by both MOLA data and the morphology of the central tract.

186 **Zone 4b:** For ~4 km from the end of Zone 4a, the PTF is partly surfaced by the continuation of  
187 FN and FS (Fig. 5b), here ~1.5 – 2.5 km-wide. They are separated by longitudinally lineated or  
188 pitted PTF, fronting the depressed, disrupted central tract of Zone 4a, and bounded laterally by  
189 the piedmont surface. FN extend from the terminal edge of the disrupted Zone 4a surface. FS  
190 are higher than FN and descend from Zone 4a onto Zone 4b, bifurcating into northern and  
191 southern branches (FSn and FSs respectively), which together surround a ~5 km-

diameter circular mass of intensely pitted (including internal subsidiary pits), hummocky material. FS cross-laps FN, passes under the northwestern limb of the circular mass but reappears at its northeastern limb. FSs curve towards the southern margin of the trench, around the perimeter of the circular mass, re-joining and cross-cutting FN. Overlying and infilling parts of FSs are discontinuous patches of low-albedo material. Where FS re-joins FN (Fig. 5b, small arrows), this material partly infills furrows in the northern region. Given the cross-cutting relationship (southern furrows postdate the northern), this could indicate that the low-albedo material was transported along the FS furrow system and backfilled FN (Fig. 5b).

Fig. 5 here

**Zone 4c:** The Zone 4b furrows terminate at a scoop-shaped embayment cutting into the PTF (Fig. 6a). Based on shadows, the exposure along the sloping backwall appears tiered, suggesting that the furrowed material is layered. The embayment is bisected by an axial trough, and narrow, sinuous ridges are present on its floor (Figs. 6a and 6b). In the proximal end of this ridge system, two parallel ridges (RpN and RpS) are partially exposed. Trending WNW-ESE, they are initially continuous for ~0.5 – 2 km, and are ~70 m wide and 150 – 300 m apart. After about 3 km, the two ridges diverge and develop subtly different morphologies. RpS has a simpler form than RpN, being sharp crested and nearly continuous, formed of only four or so sections, over a path length of ~10 km. RpN, though, becomes discontinuous and more complex, bifurcating to form a medial pair of ridges consisting of offset straight segments, 80 –

213 300 m long, forming a disjointed curve, with up to 140 m between segments. The terminus of  
214 one branch of the bifurcated RpN ridge system is obscured by rough terrain, but the southern  
215 branch is visible, although discontinuous, until it either becomes confluent with, or is cross-  
216 lapped by, the end of RpS. Beyond this confluence, the two ridge systems form a single,  
217 discontinuous, ~6 km long, segmented complex (Rd) of short, broad-crested, lozenge-like  
218 ridges, each a few hundred meters or less in length and separated longitudinally by >50 m.  
219 Some segments appear to cross-lap others, some to run beside adjacent segments.

220

221 Fig 6 here

222

### 223 3.5. Zone 5

224 East of the Zone 4c ridges, the terrain drops slightly into a flat elongated basin, 250 km long  
225 (north-south) and up to 70 km wide (Figs 2 and 7a). The margin of the basin fill has an  
226 extremely narrow elevation range (Fig 7b) but its surface is subtly fractured in places, gently  
227 hummocky and morphologically distinct from Zone 4c and the bounding piedmont zones. The  
228 basin fill surface contains large numbers of impact craters but no ring-mold forms (although  
229 many have apparently smooth, perhaps “icy”, fills) as present on the LVF. Lobate debris aprons  
230 marginally transgressing the fill in the north of the basin (centred on X in Fig. 7b) contain only a  
231 few impact craters, some being ring-mold forms.

232

### 3.6. Age of the system – impact crater size-frequency statistics

To estimate the formation and modification ages of Zone 1 and 5, the size-frequency distribution of impact craters was measured on the 180 km<sup>2</sup> surface of the LVF/GLF east of the apex and across the entire 6100 km<sup>2</sup> of Zone 5, defined by its mapped marginal contact (Fig 7b). In Zone 5, using HRSC nadir image h1423\_0001, only craters >20 pixels across (250m) were reliably identified. For Zone 1, less noisy CTX images were used (Appendix 3) and craters larger than about 15 pixels across (90m) were recorded.

Computing the crater retention age in *CraterStats*, using the Ivanov (2001) Mars production function and the Hartmann and Neukum(2001) chronology, the crater diameter range best matching an isochron for Zone 1 was 90 - 350 m (54 craters) and, for Zone 5, 300 - 2000 m (247 craters). Resultant crater retention ages for Zone 1 and Zone 5 are, respectively,  $150 \pm 20$  Ma and  $1.6 \pm 0.1$  Ga (Appendix 3, Fig A2). This is probably a very approximate LVF age: the few larger craters (or traces of craters) on the LVF might follow an older isochron, on the order of 1-1.5 Ga, which could better represent the LVF formation age, rather than its modification age. With such small numbers (2-3 craters), though, this is a very tentative conclusion. The Zone 5 crater retention age, although complicated by the small number of craters > 2 km, points to basin filling in the Amazonian. This is later than envisaged by Tanaka et al. (2014) but not inconsistent with their interpretation of the fill as volcanic.

## 4. Interpretation

The LVF (Fig. 2) comprises a complex of lobes descending from the valley sides towards its hypsometric axis, which marks the front between opposing lobes that are competing to follow

254 the same fall line during longitudinal advection downvalley. There are no individual sources  
255 along the valley sides, only scoured chutes descending from the convergent troughs incised into  
256 the plateaus overlooking the valley, indicating that the lobes originated on these plateaus, not  
257 within the valley. Moreover, the chutes are not hanging valleys but are incised into the valley  
258 walls and graded to the surface of the LVF complex. Hence, the valley was only partially filled,  
259 and likely not originally cut, by the LVF lobe-complex sourced on the plateaus.

260 The lobate, lineated morphology of Zone 1, including the presence of ring mold craters (Fig.  
261 3a) indicative of an ice-rich substrate (Kress and Head, 2008), are consistent with the LVF being  
262 a topographically-bounded icy body. Moreover, the LVF possesses the following characteristics  
263 considered to be diagnostic features of martian mid-latitude glaciers (Souness et al., 2012) and  
264 which shed more light on the formation and evolution of the entire system. (1) The LVF is  
265 surrounded by topography modified by viscous flow over or around obstacles, best represented  
266 by the confluent troughs (Fig. 2) graded to the surface of the main LVF from the adjoining  
267 upland along distinct topographic corridors (Bennett, 2003). (2) The LVF is texturally and  
268 morphologically distinct from upland summital areas and inter-valleys. (3) It displays foliation  
269 indicative of down-slope flow, especially in the form of lobes lineated by narrow ridges and  
270 furrows (Fig. 3a). Longitudinal flow lineations have been attributed to lateral compression  
271 where topography funnels ice into narrow tongues (Stokes and Clark, 1999) and experiences  
272 rapid longitudinal extension (Glasser and Gudmundsson, 2012) in a setting of long-term flow  
273 stability (Holt et al., 2013; Glasser et al., 2015), explanations clearly consistent with the context  
274 of the LVF here. (4) The LVF is a distinct, narrow flow-form laterally confined by the upland  
275 valley sides, and (5) bounded longitudinally by pitted, cross-valley, moraine-like ridges (MLR) in

276 Zone 2. MLR are indicative of the staged retreat of the LVF terminus by backwasting and  
277 marginal stagnation and are, therefore, indicative of dynamic compositional and process  
278 thresholds. (6) Throughout its course, the LVF has a viscous 'valley fill' surface.

279

280 Fig. 8 here

281

282 These six key characteristics (Souness et al., 2012) are consistent with the LVF being a  
283 topographically bounded glacier that was supplied by ice converging from surrounding upland.  
284 Hence, it is part of an assemblage of glacial forms between latitudes 30° to 60° in both  
285 hemispheres of Mars (Hubbard et al., 2014).

286 The straight, parallel-sided form of the valley, especially the narrow, trench-like form of the  
287 western branch of the system (Fig. 2), bears a striking resemblance to two trenches that cut  
288 through highlands zones well west of the Phlegra Montes (Fig. 9) and which are directly along  
289 strike (WNW-ESE). The first, ~230 km distant along strike, is ~60 km long. The second, ~450 km  
290 distant along strike, is ~110 km long. Both are about 5 km wide, the first being occupied by LVF,  
291 the second by a pitted mantle superimposed by lateral mass wasting lobes. The coherence in  
292 strike and scale of these trenches with the valley in Phlegra makes it likely that they all  
293 originated as grabens along the same fault system, during regional rifting prior to the formation  
294 of the LVF. Hence, the trunk valley occupied by the LVF is probably not a typical LVF context,  
295 being part of a very long fault system, expressed intermittently as a graben.



296

297 Fig 9 here

298

299 The cross-valley pitted ridges in Zone 2 are interpreted as recessional moraines that formed  
300 after the glacier had retreated from the piedmont. The circular pitted mass in Zone 4b is  
301 interpreted as dead-ice producing incipient ice stagnation topography. This, together with the  
302 intense pitting of the PTF in Zone 3 and the pattern of PTF disruption in Zone 4a, is evidence  
303 that the piedmont trench was formerly fully occupied by ice. The pitting of the piedmont  
304 surface, especially the proximal piedmont, is evidence that the piedmont is extensively mantled  
305 by degraded ice rich material. Hence, it is likely that the PTF is a remnant either of outlet glacial  
306 ice from the Zone 1 to 2 upland or of a piedmont LDA. The morphology of the breached  
307 bedrock ridge bounding Zone 2, including the cleaving of the originally continuous ridge into  
308 opposing headlands by incision of the breach at the lowest point of the ridge, reflect the  
309 gravitational focusing of a very narrow line of erosion along the hypsometric axis of the small  
310 valley bounded by the headlands. Bedrock breaches like this are not typical of fluvial erosion in  
311 the absence of a significant step-change in hydrology, involving either discharge (e.g. lake-  
312 bursts), baselevel lowering or cross-valley uplift. Hence, the breach is unlikely to reflect pre-  
313 glacial fluvial erosion. However, together with the presence of the sinuous re-entrant valley  
314 which appears to carve into the LVF nearly adjacent to the breached ridge, these characteristics  
315 are consistent with meltwater erosion sourced from ice immediately upslope of the breached  
316 ridge. The absence of channels up-valley of the breached ridge and the re-entrant valley

317 suggests that the meltwater was englacial, not supraglacial, although shallow supraglacial  
318 channels could have been removed by surface modification. The re-entrant does not appear to  
319 be a pre-existing fluvial or sapping bedrock channel, for there is no evidence of layering or  
320 outcrop in its flanks in its upper reaches, unlike incisions into many of the massifs surrounding  
321 the graben. It is therefore unlikely that this is a bedrock channel. However, even were this re-  
322 entrant a bedrock channel, its incision by subglacial meltwater prior to exposure by ice marginal  
323 decay could not be precluded. The sinuous axis of the re-entrant valley is evidence of proximal  
324 proglacial flow. The pitted, hummocky surface of Zone 3 and the disrupted central tract of Zone  
325 4a are consistent with the degradation of a glacial surface that experienced distal collapse along  
326 a well-defined path close to its terminus. The swaths of convex-up, undulatory forms and  
327 longitudinal furrows in Zones 4a and 4b are interpreted as fluvial sediment gravity flows. The  
328 convex-up, undulatory forms are fluvial bars, incised by multiple-channel longitudinal flows  
329 (the furrows). The sudden appearance of incised bars and furrows at the terminal edge of the  
330 central part of Zone 4a suggests that they originated from sediment-bearing liquid flows  
331 sourced in this degrading material. The longitudinally furrowed surfaces extending from the  
332 lateral parts of Zone 4a into Zone 4b suggests that the disrupted central tract of Zone 4a was  
333 the main source of liquid flows, but that some may have been sourced from the Zone 3/4a  
334 boundary. The darkest surfaces in Zone 4b and 4c probably represent distal fines from these  
335 flows. An alternative hypothesis is that these units are distal lavas, originating in the upland or in  
336 the Zone 5 basin that were emplaced before occupation by the LVF. It is unlikely that the Zone 4  
337 materials are lavas associated with volcanic filling of the Zone 5 basin because they are uphill of  
338 the basin, which is extremely flat. Also, the PTF is not characterised by flow lobes, break-outs or

evidence of fluid propagation along interior channels, arguing against emplacement from the west. Although not definitive evidence against lava, the contextual landsystem is so overwhelmingly glacial that the overarching interpretation of the PTF in Zones 3 and 4a, as a remnant glacial thermokarst assemblage, including some mantling proglacial glaciofluvial deposits, is robust even if not definitive without groundtruth. In this context, the central pitted mass (Zone 4b) is consistent with being an isolated mass of dead-ice, abandoned either during ice marginal backwasting or LDA decay. The form running transverse (Fig. 6b, X-Y) to the Zone 4c ridge system is possibly an ice margin remnant, although better resolution imagery is needed to make a more confident interpretation. However, in the context of the total landsystem, the sinuous ridges themselves are consistent with a subglacial to proglacial transition due to ice marginal decay in this location. Taken together with the rest of the landsystem, beginning in the upland at an extant but marginally decayed glacier and terminating in the piedmont at an esker-like ridge system, the relative locations and morphological characteristics of all the components of the landsystem (Fig. 8) are mutually consistent and consistent with its interpretation as a complete glacial system that experienced decay involving at least one phase of melting and significant meltwater production.

The coherence in strike and width between the valley hosting the LVF and PTF and the two grabens further east (Figs 1 and 9) indicate at least a structural control in the path of the system. However, the breached ridge and sinuous re-entrant valley in the pitted terminal zone of the LVF (Zone 2) and the glacial thermokarst and channels on the PTF are evidence of melting associated specifically with the presence of a regionally significant fault line (the fault persists along strike for 630 km). The interpretation of the LVF and PTF as a complex glacial system

361 reflecting a large degree of geological control and exhibiting evidence of meltwater activity  
362 helps to explain the sinuous ridges in Zone 4c, for their sinuous form, dimensions and context  
363 are consistent with an interpretation as eskers. The presence of eskers reflects the former  
364 extension of the glacial system to the margins of Zone 5, prior to the staged retreat of the  
365 system into the upland valley.

366 Relatively small eskers, on the scale of the Phlegra ridges, form on Earth at the terminus of  
367 glaciers when surface meltwater penetrates to the bed through moulins and crevasses (Boulton  
368 et al., 2001). It is possible that the channels in Zone 4 played a role in this respect, as hydraulic  
369 coupling between the glacier surface and base is known to amplify basal hydrostatic pressure  
370 and generate water-filled basal cavities (e.g. Boulton et al., 2001). In Phlegra, however,  
371 geothermal undermelt could have enhanced cavity production through roof melting as a  
372 consequence of basal meltwater production.

373 The continuous but diverging-converging, sinuous form and scale ( $10^1$  m-wide,  $10^3$  m-long, over  
374 an area 4.5 km by 1.25 km) of the Zone 4c ridges are analogous to many small-scale esker  
375 complexes on Earth, both recent (e.g. in Spitsbergen, Fig. 10a) and Pleistocene (e.g. the  
376 Knockbarron esker in Ireland, Fig. 10b; a complex of closely-spaced,  $10^1$  m-wide,  $10^3$  m-long  
377 sinuous ridges over an area 1.83 km by 0.63 km; and esker-net complexes in Maine, northeast  
378 USA, Fig. 10c). Continuous ridge eskers (Warren and Ashley, 1994) are tunnel fill deposits,  
379 representing sedimentation along subglacial meltwater tunnels melted upwards into overlying  
380 ice (i.e. R-channels) and bounded by a stable ice margin. However, the moderate length and  
381 directional coherence but lateral off-setting of the segments composing some of the ridges

here are consistent with a more unstable, probably crevassed, ice margin (Warren and Ashley, 1994). The continuity and sinuosity of RpS are consistent with an interpretation as continuous tunnel fill eskers, but the sharp ridge-crests are indicative of a phase of strong melting and rapid lowering of the overlying ice together with its subglacial debris load (Shreve, 1985). The segmented ridge assemblage Rd could be interpreted as remnants of a single esker (Shreve, 1985) or as a sequence of short beads (Warren and Ashley, 1994). If a short-beaded esker, it represents drainage to an aqueous ice margin characterised by short periods of stability but generally rapid retreat (Warren and Ashley, 1994). It is also possible that Rd represents a time-transgressive assemblage of segments originating first in flows along a medialRpN-conduit and then cross-lapped by RpS. If the vertical arrangement of the ridges in the distal zone represent a chronological evolution, RpS is likely to have been the last active ridge in the complex. Caution is required, however, in the absence of sedimentological exposure, in interpreting time and space relationships among these ridges.

Fig. 10here.

Given the absence of eskers elsewhere in the region, and the likelihood that the PTF is the surface expression of an underlying subglacial fault, the eskers probably reflect spatially focused melting due to enhanced heat flux along the fault strike (*cf.* Lysak, 1992; Lysak and Sherman, 2002; Clauser and Villinger, 1990; Schroeder et al., 2014). It is unlikely that icethickness alone could have caused basal melting; maximum upland ice thickness is unlikely

to have exceeded 1.5 km (based on the elevation difference between the plateau around Zone 1 and the surface of Zone 4) and was probably much less, judging by the graded chutes confluent with the LVF surface. Combined with very low mean Amazonian atmospheric temperature (Fastook et al., 2012), the resultant maximum excess pressure of ~5 MPa is insufficient for basal pressure melting. Hence, an additional heat flux was required for melting on the scale suggested by the presence of both eskers and extensive channels (cf. Fastook et al., 2012).

The orientation, structure and fault features of Phlegra Montes might have a genetic association with the Elysium Volcanic Centre (EVC; Moore, 1985). Consistent with Vaucher et al. (2009), Platz and Michael (2011) found evidence of EVC and associated regional activity (Elysium Planitia and Cerberus Fossae) spanning ~3.4 Ga to only ~1.4 Ma. Hecates Tholus, the closest EVC component to Phlegra Montes, formed ~350 Ma. The filled grabens in and east of Phlegra Montes are coherent in strike with a super-regional set of linear troughs extending across the northern flank of the Elysium Rise, through Galaxias, to Utopia Planitia. Hence, given the longevity of EVC volcanic processes and the scale of their effects, including the possible genetic linkage between the EVC and faulting in Phlegra Montes, the elevation of heat flux along the Phlegra graben system while occupied by a piedmont outlet glacier during the Late Amazonian cannot be precluded. The construction of the main valley glacier in Phlegra by the convergence of several plateau tributaries into a pre-existing strike-valley (Zone 1), could reflect enhanced geothermal heat flux along strike and the development of a positive feedback

424 involving increased basal temperature, melting, basal lubrication and flow velocity (cf. Bennett,  
425 2003). Enhanced geothermal heat flux extending from the strike valley and along the PTF could  
426 have amplified this positive feedback, with outputs as higher ice velocity and increased  
427 advection, surface down-draw and basal meltwater production.

428 As no evidence has been presented of esker formation analogous to the Phlegra Montes system  
429 elsewhere on Mars, although other fault-bounded glacial systems exist (e.g. Levy et al., 2007), it  
430 is worth speculating on the factors that could determine esker potential as an epiphenomenon  
431 of enhanced geothermal heatflux. For this to occur, a glacier must occupy a fault during a  
432 phase, and at a location, characterised by sufficiently elevated heat flux, either associated with  
433 fault emplacement (with significant heat propagation lags) or reactivation. In systems like the  
434 Rheingraben or the Kenya Rift, enhanced heatflux commonly persists for  $10^7$  years, with  $10^6$  -  
435  $10^7$  years required for heat propagation from the Moho to the surface (e.g. Wheildon et al.,  
436 1994).

437 The glacial hydrological system in Phlegra is unlikely to be replicated in regions with no  
438 evidence of Late Amazonian volcanism and fault (re)activation coincident with glaciation.  
439 Moreover, because enhanced heatflux is highly concentrated but variable along rift axes,  
440 consequential meltwater production and esker formation might have occurred only in  
441 restricted spatial and temporal contexts within a single bounding fault, even in the Phlegra  
442 case. Considering the Dorsa Argentea eskers, perhaps indicative of Noachian-Hesperian warm-  
443 based glacial conditions, Fastook et al. (2012) concluded that atmospheric temperatures must  
444 reach  $-75$  to  $-50^\circ\text{C}$  for significant basal melting to occur in settings characterised by typical

geothermal heat fluxes (45-65 mW.m<sup>-2</sup>). This represents an enormous climate change but still results in cold based glaciers, implying that geothermal conditioning is strongly implicated in the production of martian eskers. If subglacial water flux generally tends to be low, due to inherently low atmospheric temperatures, subcritical geothermal heat flux and weak melting, the widespread development of R-channels required for esker formation is, most likely, precluded. Only beneath ice experiencing enhanced melting, evidently due to significant atmospheric temperature excursions and/or elevated geothermal heat flux (Fastook et al., 2012), could subglacial water flux increase both down-glacier and towards the glacier bed sufficiently to allow R-channels to form and remain open. Hence, on Mars, the development and survival of R-channels capable of transporting sufficient quantities of both meltwater and sediment required for esker formation might be very rare, reflecting the short time periods and limited locations in which both atmospheric temperatures and geothermal heat flux combine to exceed the required critical threshold. We note that channels, and other evidence of subglacial to proglacial meltwater routing, occur at the Zone 2 margin and on the Zone 4a surface of the Phlegra system. On this basis, observations of channels closely associated with glacial margins elsewhere on Mars should be re-examined as possible evidence of subglacial to proglacial meltwater routing.

Fig 11 here

## **5. Implications of glacial melting and liquid flows**



## 5.1 Surface flows

Hubbard et al. 2014 concluded that the extremely rare evidence of supraglacial melting on Mars points only to short-lived, unorganized liquid flows and that no evidence exists of proglacial fluvial activity. However, Fassett et al (2010) described glaciers that showed evidence of limited surface melting and proglacial drainage in the Amazonian. Consequently, the consensus is that the thermal regimes of extant martian glacier-like forms are cold, although perhaps not always in the past (Hubbard et al., 2011). The Phlegra glacial system (upland and piedmont) is important, therefore, for it shows evidence of ice-contact glaciofluvial breaching of the ridge and incision of the re-entrant in the terminal zone of the LVF (Zone 2) and suggests that water from Zone 2 was exported to the PTF in Zone 3. The fluvial systems in Zones 4a and 4b probably originated from sediment-bearing liquid flows sourced on the Zone 3 PTF and from the longitudinally furrowed surface extending beyond the disrupted surface in Zone 4a.

## 5.2 Basal flows

Interpretations of sinuous ridges on Mars as eskers include the implication that extensive, thick wet based glaciation has occurred (e.g. Kargel and Strom, 1991; Banks and Pelletier, 2008; Banks et al., 2009; Head, 2000; Bernhardt et al., 2013). However, because none of these eskers is associated with an intact glacier, very little is known about the possible range of subglacial hydrology associated with martian esker formation. In the Phlegra landsystem, while the morphology and dimensions of the sinuous ridges in Zone 4c are consistent with their interpretation as eskers, the PTF shows the direct evidence of both subglacial and supraglacial flows in the same glacial system, although the supraglacial flows appear to originate from the

emergence of englacial flows(e.g. in Zone 4a from probable glacial ice in Zone 3). Similar emergences from englacial channels are common in glaciers in geothermal zones on Earth (e.g. Waltham, 2001), including the development of sub-aerial channels on surfaces exposed due to the thinning and collapse of ice roofs.

Generally, the presence of eskers implies significant basal routing of meltwater, but not necessarily the production of meltwater at the base. On Earth, subglacial meltwater flows often reflect hydraulic coupling between the glacial surface and the base. Englacial to subglacial conduits that develop within hydraulically coupled systems represent tunnel-confined basal flows of water fed from melting, flows and standing water at the glacier surface (Benn et al., 2012). In Phlegra, however, it is likely that the eskers and surface channels carved by emergent englacial flows reflect conductance of geothermal heat from the subglacial graben through the ice and, therefore, that the formation of the eskers involved both basal production and routing of meltwater. The cessation of excess heat flux from the graben and, therefore, transition of the overlying glacier to a cold based thermal regime in equilibrium with prevailing climate, probably explains the survival of the system, in company with the surrounding glaciers. It should be noted, though, that even on Earth, eskers can be preserved as glacial retreat strands them in an evolving proglacial outwash system (e.g., Fig. 10a).

## **6. Conclusions**

The assemblage of landforms described here is consilient in relief, relative topography and morphology with an interpretation as a wet-based glacial system (Figs 8 and 11). This system

presents what appears to be the first identification of martian eskers that can be directly linked to their parent glacier. The observations demonstrate the presence of a wet-based system, implying that, where there is sufficient heat, glaciers on Mars will attain warm/wet based regimes. However, here the energy required for melting came from bottom-up geothermal heating, rather than being due to changes in climate. In the absence of such geological conditioning, glaciers on Mars seem largely incapable of achieving these regimes, overarching climatic control producing only cold based glaciers. Regarding the Phlegra Montes landsystem, it remains to be definitively established if the graben was thermally active during the glacial period, or simply a pre-existing topographic corridor or sink. If it was active, could glacial fast-flow have been triggered along the geothermal-topographic corridor, enhanced geothermal heating creating a positive feedback involving melting and glacio-dynamics that led to the formation of this possibly unique system? Resolving these problems and effectively contextualising the answers will require searching elsewhere on Mars for evidence of geothermal influences on lowland glaciers, including supraglacial melting and proglacial discharges, in the absence of direct observations of eskers.

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## REFERENCES

- Banks, M. E. and Pelletier, J. D. (2008). Forward modeling of ice topography on Mars to infer basal shear stress conditions, *J. Geophys. Res.*, 113, E1, <http://dx.doi.org/10.1029/2007JE002895>.
- Banks, M. E., N. P. Lang, J. S. Kargel, A. S. McEwen, V. R. Baker, J. A. Grant, J. D. Pelletier, and R. G. Strom (2009), An analysis of sinuous ridges in the southern Argyre Planitia, Mars using HiRISE and CTX images and MOLA data, *J. Geophys. Res.*, 114, E09003, doi:10.1029/2008JE003244.
- Benn, D.I., Bolch, T, Hands, K., Gulley, J., Luckman, A., Nicholson, L.I., Quincey, D., Thompson, S., Toumi, R. and Wiseman, S. (2012) Response of debris-covered glaciers in the Mount Everest region to recent warming, and implications for outburst flood hazards, *Earth-Science Reviews*, 114, 156–174.
- Benn, D., Gulley, J., Luckman, A., Adamek, A. and Glowacki, P.S. (2009). Englacial drainage systems formed by hydrologically driven crevasse propagation, *Journal of Glaciology*, 55, 191
- Bennett, M.R. (2003). Ice streams as the arteries of an ice sheet: their mechanics, stability and significance, *Earth-Science Reviews*, 61, 309 – 339. doi:10.1016/S0012-8252(02)00130-7.
- Bernhardt, H., Hiesinger, H., Reiss, D., Ivanov, M., and Erkeling, G. (2013). Putative eskers and new insights into glacio-fluvial depositional settings in southern Argyre Planitia, Mars, *Planet. Space Sci.*, 85, 261–278.

552 Boulton, G.S., Dongelmans, P., Punkari, M. and Broadgate, M. (2001). Palaeoglaciology of an ice  
 553 sheet through a glacial cycle: the European ice sheet through the Weichselian, *Quaternary*  
 554 *Science Reviews*, 20, 591-625.

555 Boulton, G. (2010). Drainage pathways beneath ice sheets and their implications for ice sheet  
 556 form and flow: the example of the British Ice Sheet during the Last Glacial Maximum, *J.*  
 557 *Quaternary Sci.*, 25, 483–500.

558 Clauser, C. and H. Villinger, H. (1990). Analysis of conductive and convective heat transfer in a  
 559 sedimentary basin, demonstrated for the Rheingraben, *Geophys. J. Int.*, 100, 393-414.

560 Dickson, J.L., Head, J.W., Marchant, D.R. (2010). Kilometer-thick ice accumulation and glaciation  
 561 in the northern mid-latitudes of Mars: Evidence for crater-filling events in the Late  
 562 Amazonian at the Phlegra Montes, *Earth and Planetary Science Letters*, 294, 332–342.

563 Fassett, C.I., Dickson, J.L., Head, J.W., Levy, J.S and Marchant, D.R. (2010). Supraglacial and  
 564 proglacial valleys on Amazonian Mars, *Icarus*, 208, 86–100.

565 Fassett, C. I., Levy, J.S., Dickson, J.L. and Head, J.W. (2014). An extended period of episodic  
 566 northern mid-latitude glaciation on Mars during the Middle to Late Amazonian: Implications  
 567 for long-term obliquity history, *Geology*, 42 (9), 763–766, doi: 10.1130/G35798.1.

568 Fastook, J.L., Head, J.W., Marchant, D.R., Forget, F., Madeleine, J.-B. (2012). Early Mars climate  
 569 near the Noachian-Hesperian boundary: Independent evidence for cold conditions from  
 570 basal melting of the south polar ice sheet (Dorsa Argentea Formation) and implications for  
 571 valley network formation. *Icarus*, 219, 25–40. doi:10.1016/j.icarus.2012.02.013.

572 Fastook, J.L., Head, J.W. and Marchant, D.R. (2014). Formation of lobate debris aprons on Mars:  
 573 Assessment of regional ice sheet collapse and debris-cover armoring, *Icarus*, 228, 54–63.

574 Glasser, N.F. and Gudmundsson, G.H. (2012). Longitudinal surface structures (flowstripes) on  
 575 Antarctic glaciers, *The Cryosphere*, 6, 383–391, 2012 [www.the-cryosphere.net/6/383/2012/](http://www.the-cryosphere.net/6/383/2012/)  
 576 doi:10.5194/tc-6-383-2012.

577 Glasser, N.F. Jennings, S.J.A., Hambrey, M.J. and Hubbard, B. (2015). Origin and dynamic  
 578 significance of longitudinal structures (“flow stripes”) in the Antarctic Ice Sheet, *Earth*  
 579 *Surf. Dynam.*, 3, 239–249, [www.earth-surf-dynam.net/3/239/2015/](http://www.earth-surf-dynam.net/3/239/2015/) doi:10.5194/esurf-3-239-  
 580 2015.

581 Hartmann, W. K., and G. Neukum (2001), Cratering Chronology and the Evolution of Mars,  
 582 *Space Sci Rev*, 96(1-4), 165–194.

583 Head, J. W. (2000). Tests for ancient polar deposits on Mars: Origin of esker-like sinuous ridges  
 584 (Dorsa Argentea) using MOLA data, *Lunar and Planetary Science Conference, XXXI*, Abstract  
 585 1116.

586 Holt, J.W., Safaeinili, A., Plaut, J.J., Head, J.W., Phillips, R.J., Seu, R., Kempf, S.D., Choudhary, P.,  
 587 Young, D.A., Putzig, N.E., Biccari, D. and Gim, Y. (2008). Radar Sounding Evidence for Buried  
 588 Glaciers in the Southern Mid-Latitudes of Mars, *Science*, 322, 1235 – 1238,  
 589 doi:10.1126/science.1164246.

590 Holt, T.O., Glasser, N.F., Quincey, D.J. and Siegfried, M.R. (2013). Speedup and fracturing of  
 591 George VI Ice Shelf, Antarctic Peninsula, *The Cryosphere*, 7, 797–816, 2013 [www.the-](http://www.the-cryosphere.net/7/797/2013/)  
 592 [cryosphere.net/7/797/2013/](http://www.the-cryosphere.net/7/797/2013/) doi:10.5194/tc-7-797-2013).

593 Hubbard, B. and Nienow, P. (1997). Alpine subglacial hydrology, *Quaternary Science Reviews*  
 594 16(9), 939-955.

595 Hubbard, B., Milliken, R.E., Kargel, J.S., Limaye, A., Souness, C. (2011). Geomorphological  
 596 characterization and interpretation of a mid-latitude glacier-like form: Hellas Planitia, Mars,  
 597 *Icarus*, 211, 330–346.

598 Hubbard, B., Souness, C. and Brough, S. (2014). Glacier-like forms on Mars, *The Cryosphere*, 8,  
 599 2047–2061, [www.the-cryosphere.net/8/2047/2014/](http://www.the-cryosphere.net/8/2047/2014/), doi:10.5194/tc-8-2047-2014.

600 Ivanov, B. A. (2001), Mars/Moon Cratering Rate Ratio Estimates, in *Chronology and Evolution of*  
 601 *Mars*, vol. 12, edited by R. Kallenbach, J. Geiss, and W. K. Hartmann, pp. 87–104, Springer  
 602 Netherlands, Dordrecht.

603 Kargel, J.S. and Strom, R.G. (1991). Terrestrial glacial eskers: analogs for martian sinuous ridges,  
 604 *Lunar and Planetary Science Conference*, XXII, 683-684.

605 Kochel, R.C. and Peake, R.T. (1984). Quantification of waste morphology in Martian fretted  
 606 terrain. In: Proc. Lunar Sci. Conf. 15. *J. Geophys. Res. Suppl.* 89, C336–C350.

607 Kneissl, T., S. van Gasselt, and G. Neukum (2011), Map-projection-independent crater size-  
 608 frequency determination in GIS environments—New software tool for ArcGIS, *Planet. Space*  
 609 *Sci.*, 59(11-12), 1243–1254, doi:10.1016/j.pss.2010.03.015.



610 Kress, A.M. and Head, J.W. (2008). Ring-mold craters in lineated valley fill and lobate debris  
611 aprons on Mars: evidence for subsurface glacial ice, *Geophys. Res. Lett.*, 35 (23), L23206,  
612 <http://dx.doi.org/10.1029/2008GL035501>.

613 Kress, A., Head, J.W., Safaeinili, A., Holt, J., Plaut, J., Posiolova, L., Phillips, R., Seu, R. and the  
614 SHARAD team (2010). Age and stratigraphic relationships in massif-debris-apron terrain in  
615 western Phlegra Montes, Mars, *41st Lunar and Planetary Science Conference*, Abstract 1166.

616 Levy, J.S., Head, J.W., Marchant, D.R., 2007. Lineated valley fill and lobate debris apron  
617 stratigraphy in Nilosyrtis Mensae, Mars: Evidence for phases of glacial modification of the  
618 dichotomy boundary, *Journal of Geophysical Research: Earth Surface*, 112, doi:10.1029–  
619 2006JE002852. doi:10.1029/2006JE002852.

620 Lysak S. (1992). Heat flow variations in continental rifts, *Tectonophysics*, 208, 309–323.

621 Lysak, S.V. and Sherman, S.I. (2002). Terrestrial heat flow in areas of dynamic influence of faults  
622 in the Baikal rift zone, *EGU Stephan Mueller Special Publication Series*, 2, 153–160.

623 Malin, M.C., Bell, J.F., Cantor, B.A., Caplinger, M.A., Calvin, W.M., Clancy, R.T., Edgett, K.S.,  
624 Edwards, L., Haberle, R.M., James, P.B., Lee, S.W., Ravine, M.A., Thomas, P.C. and Wolff, M.J.  
625 (2007). Context Camera Investigation on board the Mars Reconnaissance Orbiter, *J. Geophys.*  
626 *Res.*, 112, E5, <http://dx.doi.org/10.1029/2006JE002808>.

627 Michael, G.G. and Neukum, G. (2010). Planetary surface dating from crater size–frequency  
628 distribution measurements: Partial resurfacing events and statistical age uncertainty. *Earth*  
629 *Planet. Sci. Lett.*, 294, 223–229. doi:10.1016/j.epsl.2009.12.041.

630 Milliken, R. E., Mustard, J.F, and Goldsby, D.L. (2003). Viscous flow features on the surface of  
631 Mars: Observations from high resolution Mars Orbiter Camera (MOC) images, *J.Geophys.*  
632 *Res.*, 108(E6), 5057, doi:10.1029/2002JE002005.

633 Moore, J.M. (1985). The origin of the Phlegra Montes, Mars, Lunar and Planetary Science XVI,  
634 573-574, Abstract 16\_573M.

635 Morgan, G.A., Head, J.W. III and Marchant, D.R. (2009)Lineated valley fill (LVF) and lobate  
636 debris aprons (LDA) in the Deuteronilus Mensae northern dichotomy boundary region, Mars:  
637 Constraints on the extent, age and episodicity of Amazonian glacial events, *Icarus*, 202, 22-  
638 38.

639 Neukum, G. and Jaumann, R. (2004). HRSC: The High Resolution Stereo Camera of Mars  
640 Express. In *Mars Express: The Scientific Payload*, edited by A. Wilson, 17–35. Noordwijk: ESA  
641 Publications Division.

642 Parsons, R.A., Nimmo, F. and Miyamoto, H. (2011). Constraints on martian lobate debris apron  
643 evolution and rheology from numerical modeling of ice flow, *Icarus*, 214, 246–257.

644 Platz, T. and Michael, G. (2011).Eruption history of the Elysium Volcanic Centre, Mars,  
645 Geophysical Research Abstracts, 13, EGU2011-13166, 2011.

646 Safaeinili, A., Holt, J., Plaut, J., Posiolova, L., Phillips, R., Head, J.W., Seu, R. and the SHARAD  
647 team (2009). New radar evidence for glaciers in Mars Phlegra Montes region, *40th Lunar and*  
648 *Planetary Science Conference*, Abstract 1988.

649 Schroeder, D.M., Blankenship, D.D., Young, D.A. and Quartini, E. (2014). Evidence for elevated  
 650 and spatially variable geothermal flux beneath the West Antarctic Ice Sheet, *Proceedings of*  
 651 *the National Academy of Sciences*, 111, 25, 9070-9072.

652 Seu, R., Phillips, R.J., Biccari, D., Orosei, R., Masdea, A., Picardi, G., Safaeinili, A., Campbell, B.A.,  
 653 Plaut, J.J., Marinangeli, L., Smrekar, S.E. and Nunes, D.C., (2007). SHARAD sounding radar on  
 654 the Mars Reconnaissance Orbiter, *J. Geophys. Res.*, 112, E5,  
 655 <http://dx.doi.org/10.1029/2006JE002745>

656 Shreve, R.L. (1985). Esker Characteristics in terms of glacier physics, Katahdin esker system,  
 657 Maine, *Geological Society of America Bulletin*, 96, 639–646, doi: 10.1130/ 0016-  
 658 7606(1985)962.0.CO;2.

659 Souness, C., Hubbard, B., Milliken, R.E. and Quincey, D. (2012). An inventory and population-  
 660 scale analysis of martian glacier-like forms, *Icarus*, 217, 243–255.

661 Squyres, S W. (1979). The distribution of lobate debris aprons and similar flows on  
 662 Mars. *J. Geophys. Res.*, 84, 8087-8096.

663 Stokes, C.R. and Clark, C.D. (1999). Geomorphological criteria for identifying Pleistocene ice  
 664 streams. *Annals of Glaciology*, 28(1), 67- 74, doi:10.3189/172756499781821625.

665 Tanaka, K.L., Skinner, J.A., Dohm, J., Irwin, R.P., III, Kolb, E.J., Fortezzo, C.M., Platz, T., Michael,  
 666 G.G., Hare, T.M., 2014. Geologic map of Mars: U.S. Geological Survey Scientific Investigation  
 667 Maps, Map 3292.

668 Thompson, W. B. (2014). Maine's Eskers, Maine Geological Survey Website, Geological Site of  
669 the Month, January 2014.<http://www.maine.gov/dacf/mgs/explore/surficial/sites/jan14.pdf>,  
670 accessed on 23 April 2015.

671 Vaucher, J., Baratoux, D., Toplis, M.J., Pinet, P., Mangold, N., Kurita, K. (2009). The  
672 morphologies of volcanic landforms at Central Elysium Planitia: Evidence for recent and fluid  
673 lavas on Mars, *Icarus* 200, 39–51.

674 Waltham, T. (2001). Guide to the volcanoes of southern Kamchatka, Russia, *Proceedings of the*  
675 *Geologists' Association*, 112, 67-78.

676 Warner, N.H., Gupta, S., Calef, F., Grindrod, P., Boll, N., and Goddard, K. (2015). Minimum  
677 effective area for high resolution crater counting of martian terrains. *Icarus*, 245, 198-240,  
678 doi:10.1016/j.icarus.2014.09.024

679 Warren, W.P. and Ashley, G.M. (1994). Origins of the ice-contact stratified ridges (eskers) of  
680 Ireland, *J. Sedimentary Research*, A64, 433-449.

681 Wheildon, J., Morgan, P., Williamson, K.H., Evans, T.R. and Swanberg, C.A. (1994). Heat flow in  
682 the Kenya rift zone, *Tectonophysics*, 236, 1–4, 131-149, doi:10.1016/0040-1951(94)90173-2.

683 Zuber, M.T., Smith, D.E., Solomon, S.C., Muhleman, D.O., Head, J.W., Garvin, J.B., Abshire, J.B.  
684 and Bufton, J.L. (1992). The Mars Observer Laser Altimeter Investigation, *J. Geophys. Res.*, 97,  
685 7781–97.

686

687 Figure captions

Figure 1. Location of southern Phlegra Montes. Background image is THEMIS daytime mosaic overlain by colourised MOLA topography data (purple is low elevation, brown is high). Inset shows position of this figure in relation to a global MOLA hillshade map of Mars. Locations of other figures shown by white boxes. Image credit NASA/JPL/ASU/MOLA science team.

Figure 2. Southern Phlegra study region. a) 6m/pixel CTX mosaic of the study region. The different zones in the system are marked, as are the inferred flow directions of the lineated valley fill (LVF) occupying the east-west trending valley. b) Colourised topography (MOLA data overlain on CTX mosaic) information for the study region. Position of topographic profile (bottom panel) shown by heavy black line. Note the LVF summital region and western piedmont upland, the steep convex transition from zone 2 to 3, and the gentle slopes in zones 3-5. Image credits NASA/JPL/MSSS/MOLA science team. North is up in this and all other images unless stated otherwise.

Figure 3. LVF morphology and confluent chutes. a) Zone 1 morphology. The valley is occupied by the LVF feature. Lobate forms are picked out in white. Note also the longitudinal lineations, ring-mold impact craters (e.g., black arrow) and one fresh impact crater (white arrow). The pitted region marking the start of zone 2 is seen at the right of the image. b) Confluent valleys and chutes on the northern valley wall (arrowed). The floors of these valleys have textures similar to that seen on the LVF, and are clearly indicative of headward erosion orthogonal to

708 the main valley trend. Note that the valleys floors are topographically above the surface of the  
709 main valley LVF but connected with it via graded chutes. Image credits NASA/JPL/MSSS.

710

711 Figure 4. Zones 2 and 3. a) The thinning fill of Zone 2 is dominated by two cross-valley bands of  
712 hectometer-scale pits (labelled and arrowed) enclosed by an arcuate bedrock ridge, breached  
713 at its hypsometric axis (Q) between two headlands (A and B). The pitted fill enclosed by the  
714 ridge is indented by a narrow valley (P) and the LVF is incised by a re-entrant valley (X) that cuts  
715 through the southern limb of the bedrock ridge (south of B) and terminates in a complex of  
716 large hummocks (Y). (b) Zone 3 (the proximal piedmont) is dominated by the pitted piedmont  
717 trench fill (PTF) confined by steep lateral edges, rising up to 100 m above the fill. Extending for  
718 ~2.5 km from the ridge-breach in Zone 2 is a sinuous, thread-like crease (black arrows). The  
719 edges of the trench in this zone are indented by erosional alcoves (white arrows). Image credits  
720 NASA/JPL/MSSS.

721

722 Figure 5. Zone 4. a) Zone 4a. Note the central disrupted reach, and the two bounding, subtly-  
723 furrowed platforms (FN, north, and FS, south). b) Zone 4b. The northern and southern furrow  
724 patterns that begin in zone 4a show more organization in zone 4b. The furrows link to form a  
725 continuous, bifurcating channel-like system that is directed to the north and south of a  
726 hummocky and pitted central mass. Inferred flow directions are shown by the larger white  
727 arrows. Note the darker regions overlying parts of the southern branch and possible infilling of

728 the northern branch by this dark material – smaller white arrows with ‘?’ – presumably having  
729 been transported along the southern branch of the system. Image credits NASA/JPL/MSSS.

730

731 Figure 6. Zone 4c. a). Zone 4c contains a distinctive sinuous ridge system (box shows location of  
732 Fig. 6b) within a parabolic embayment that has tiered alcoved margins.. The linked systems of  
733 furrows, visible in zone 4a and 4b appear to terminate at the scarp that defines this parabolic  
734 embayment. b) Close-up showing details of the sinuous ridges. The pattern has been picked-out  
735 in black to show the difference in form between the northern (RpN) and southern (RpS) ridges.  
736 The form running transverse (X-Y) to the ridge system is possibly an ice margin remnant. Image  
737 credits NASA/JPL/MSSS.

738

739 Figure 7. Zone 5. a) Surface textures in Zone 5. Note the rectilinear patterns on the Zone 5  
740 surface, and patterns of what appear to be subtle fractures. Arrow shows ridge system seen in  
741 Zone 4c. b) MOLA topographic data for Zone 5. The yellow line shows the morphological  
742 boundary of zone 5. The different coloured regions indicate the areas of different elevation in  
743 the lowest parts of the region, based on MOLA gridded data. Regions higher than -3240m  
744 elevation are left uncoloured. The area characterised by marginal LDAs, in the northern part of  
745 the basin, is marked X. The background is a HRSC nadir-looking visible image. Although the  
746 basin is > 200 km in length, the variations in depth across it are only about 100m. Image credits  
747 NASA/JPL/MSSS/ MOLA science team and ESA/DLR/FU Berlin.

748

749 Figure 8. Cartogram derived from CTX imaging data (see online Appendix for image details)  
750 depicting the relative locations and major morphological characteristics of the landsystem.

751

752 Figure 9. Grabens along strike of the valley in the Phlegra study area. Locations of these figures  
753 are given in Fig.1. Image is THEMIS daytime mosaic. Image credits NASA/JPL/ASU.

754

755 Figure 10. Terrestrial analogue eskers. a) Esker system in Svalbard. These eskers have been  
756 revealed by retreat of the Austre Torellbreen glacier in southwest Svalbard. The glacier itself is  
757 just to the north of the image, with the previous glacial advance direction (inferred from  
758 lineaments, furrows and moraines) shown by the large white arrow. South of the smaller white  
759 arrows, the esker system is well-organised, with clearly defined ridges, similar in morphology to  
760 those seen in the Phlegra Montes region on Mars. Image centered at approximately 77.14N,  
761 15.18E. Image credit Norwegian Polar Institute. b) High resolution (0.5 m) aerial photograph of  
762 the Knockbarron esker near Kinnitty, County Offaly, Ireland. This esker complex consists of  
763 ridges of coarse gravels and sands deposited by subglacial meltwaters flowing locally SW-NE in  
764 subglacial tunnels and discharging into a small ice marginal lake (not shown) after the Last  
765 Glacial Maximum. Image credit DigitalGlobe. c) High-resolution (2 m) LIDAR shaded-relief image  
766 of the Monroe esker network in Maine, USA. This small section of branching, complex ridges  
767 (an 'esker-net') is part of a larger system (Thompson, 2014) and similar in morphology and



768 ridge-plan to the martian example described here. North is to the bottom in this image. Image  
769 credit Maine Geological Survey.

770

771 Fig 11. Oblique view of the entire system, created using a 6 m/pixel CTX mosaic draped over 50  
772 m grid HRSC topographic data. This viewpoint shows the continuity of the system, from upland  
773 to distal piedmont, including the eskers. The system is zoned by the prevalence of different  
774 landforms but the landforms comprising each zone are consistent with analogue glacial systems  
775 on Earth. This view also shows the relative relief of the components of the system, emphasizing  
776 both the continuity of the valley-trench lineament and the outflow of the glacier from the  
777 upland valley into and through the piedmont trench/graben. The baselevel-like nature of Zone  
778 5 is also clearly expressed.

## **Eskers in a complete, wet-based glacial system in the Phlegra Montes region, Mars**

Colman Gallagher and Matt Balme

### **Keywords (to be included in main text):**

Mars; glacier; eskers; wet base; geothermal control

### **Highlights**

- The first identification of martian eskers directly linked to their parent glacier.
- The Eskers are at the degraded terminus of a graben-confined glacier.
- The eskers are evidence of glacial melting and a wet-based regime.
- Melting was due to enhanced geothermal heat flux, not climate warming.
- These are new insights into glacial behaviour and meltwater production on Mars.

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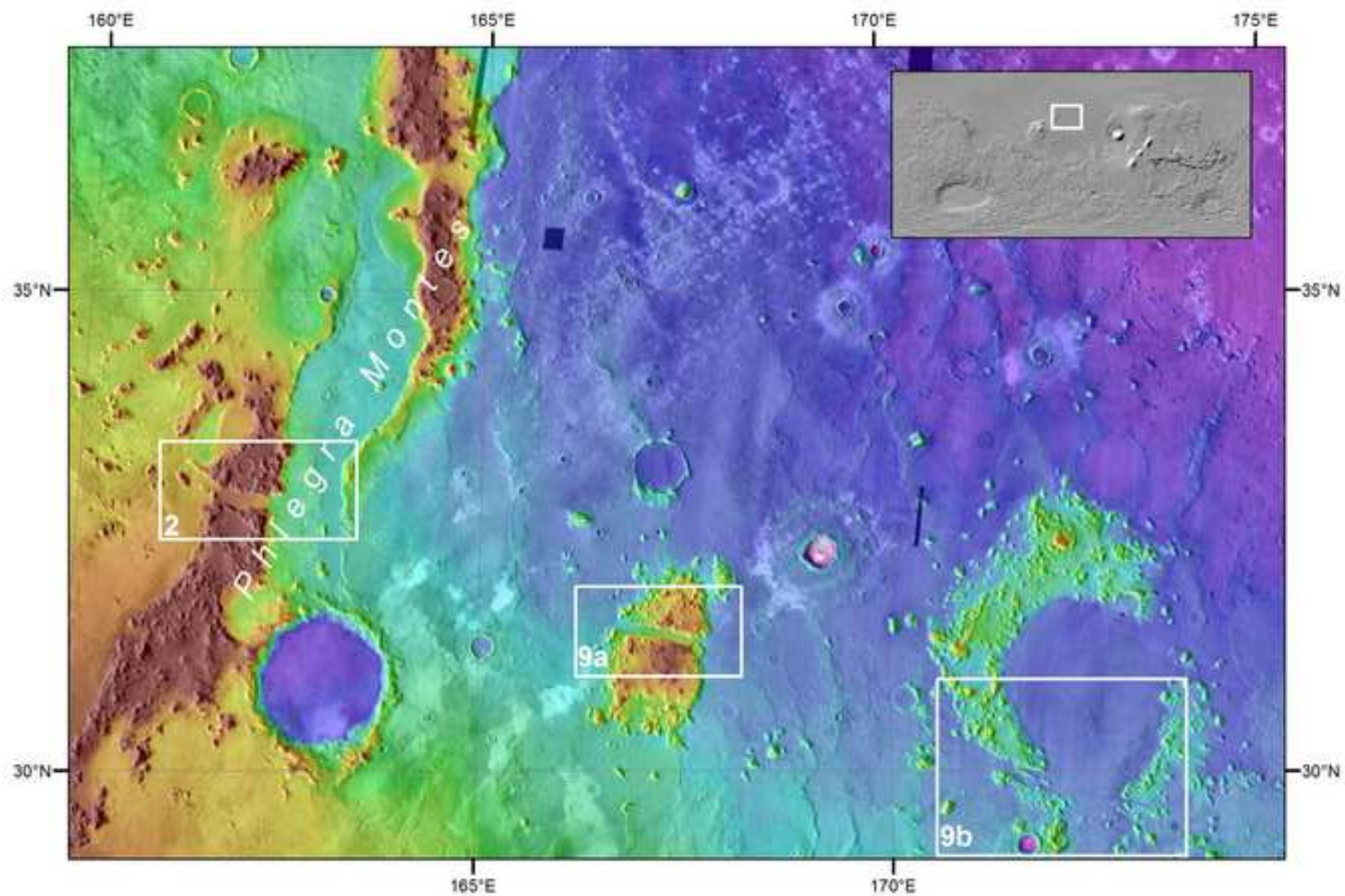


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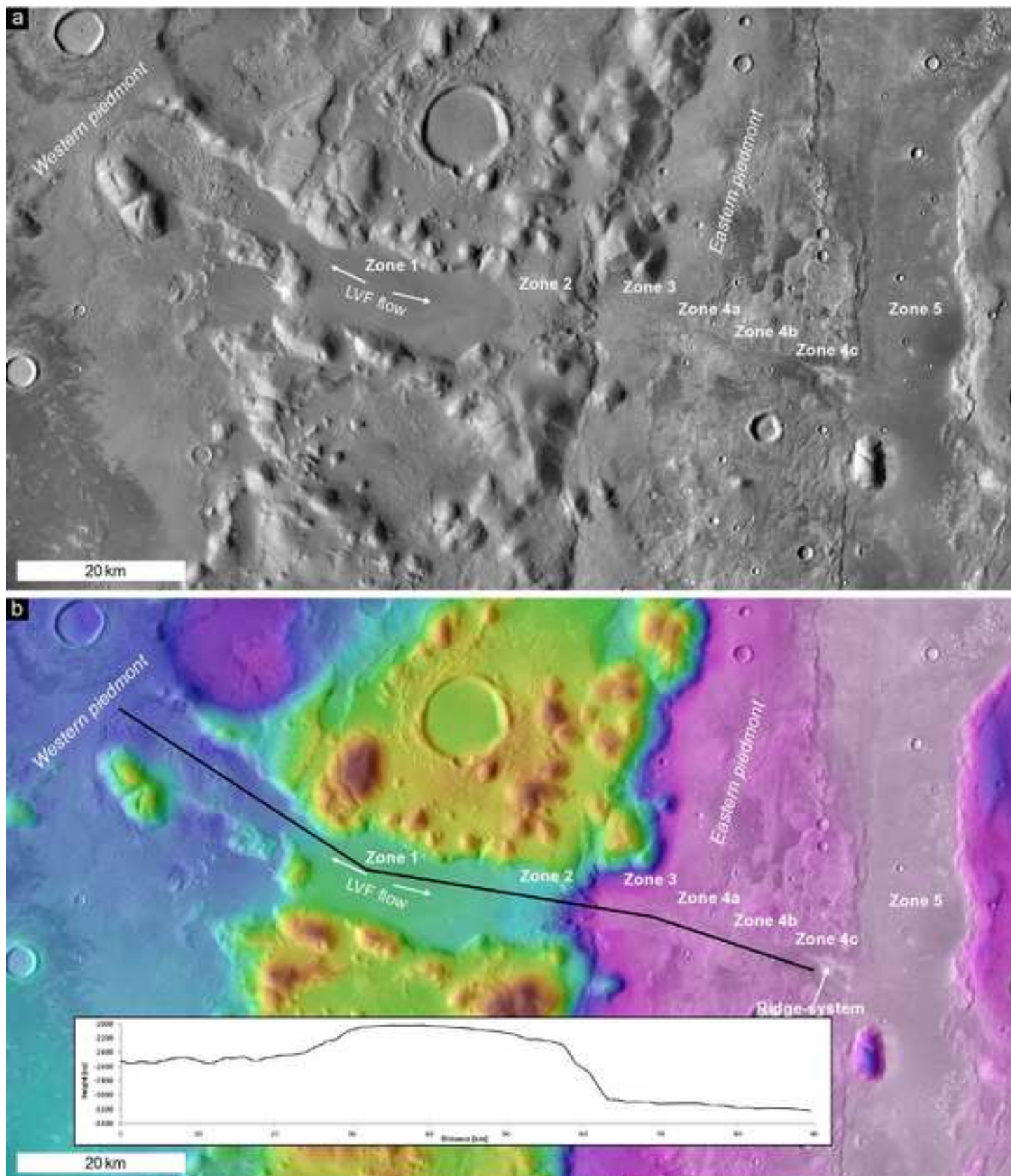




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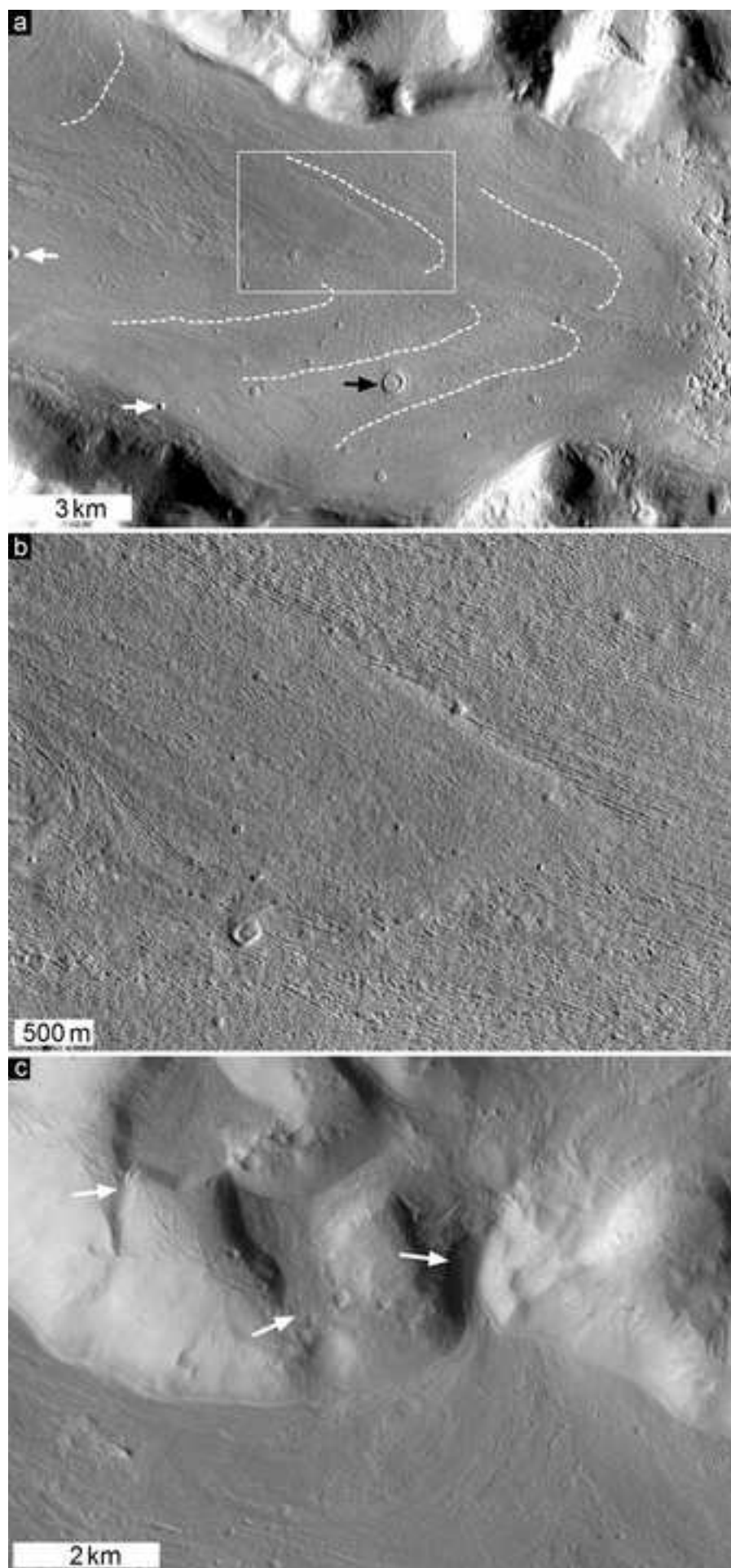


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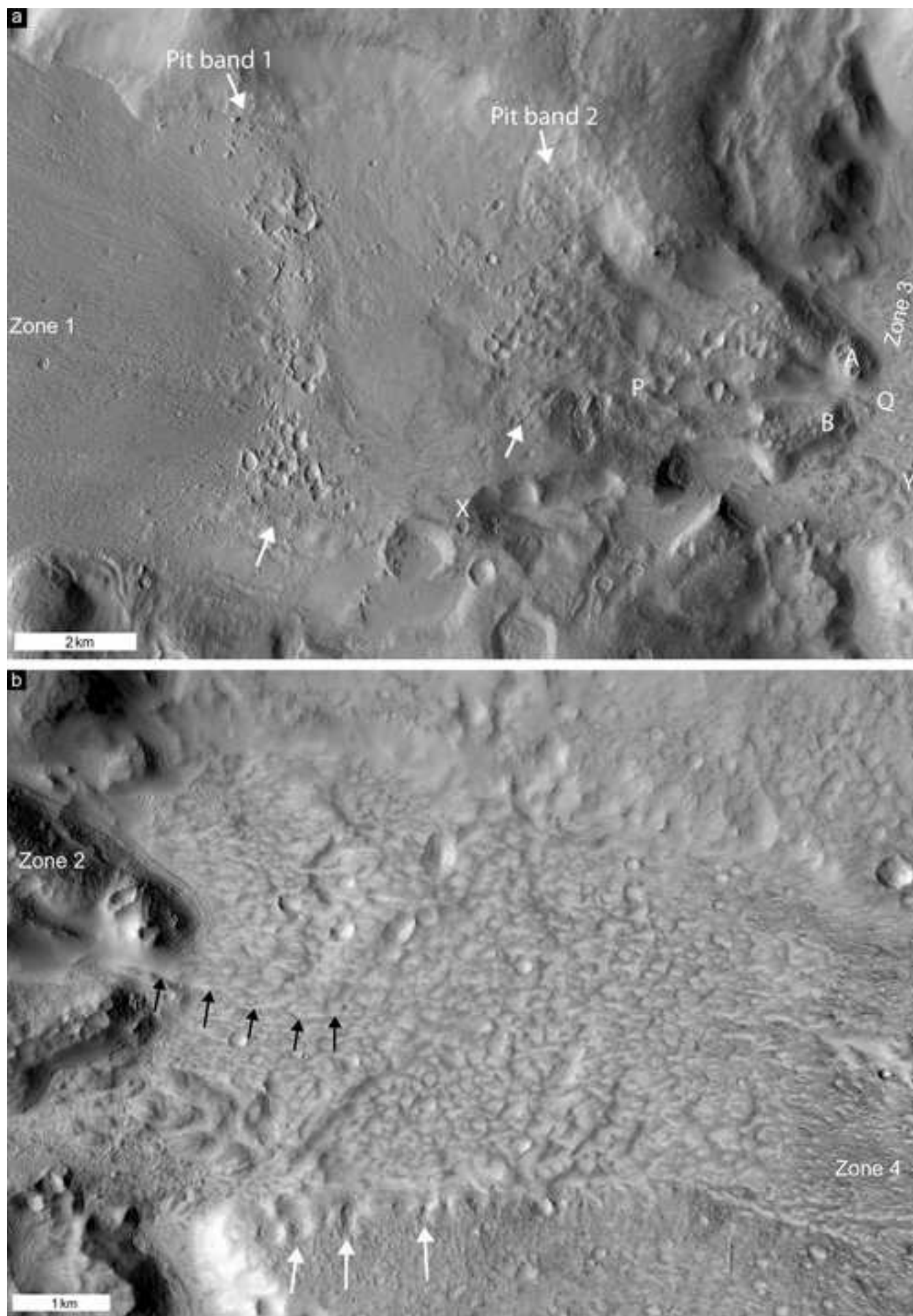




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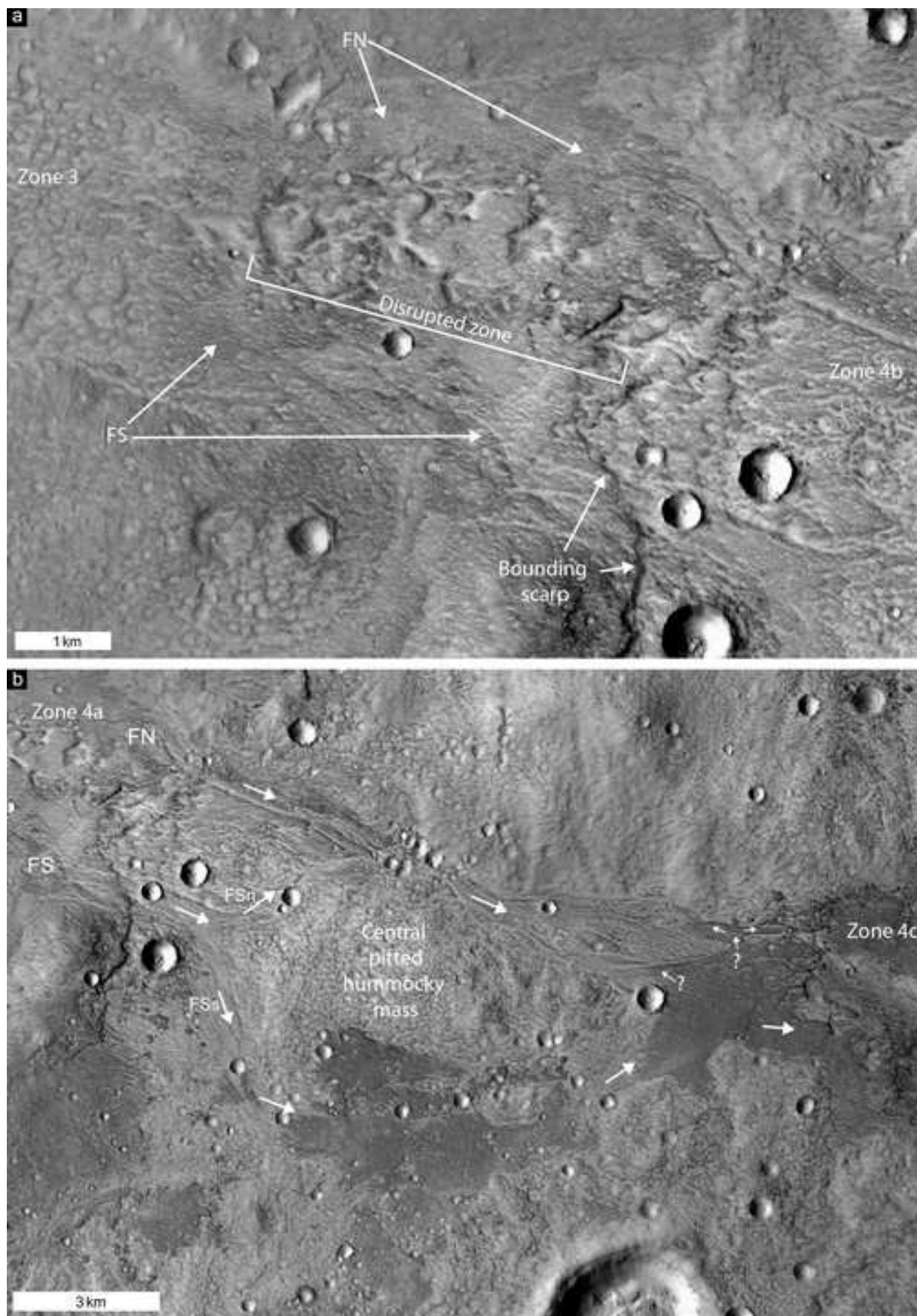
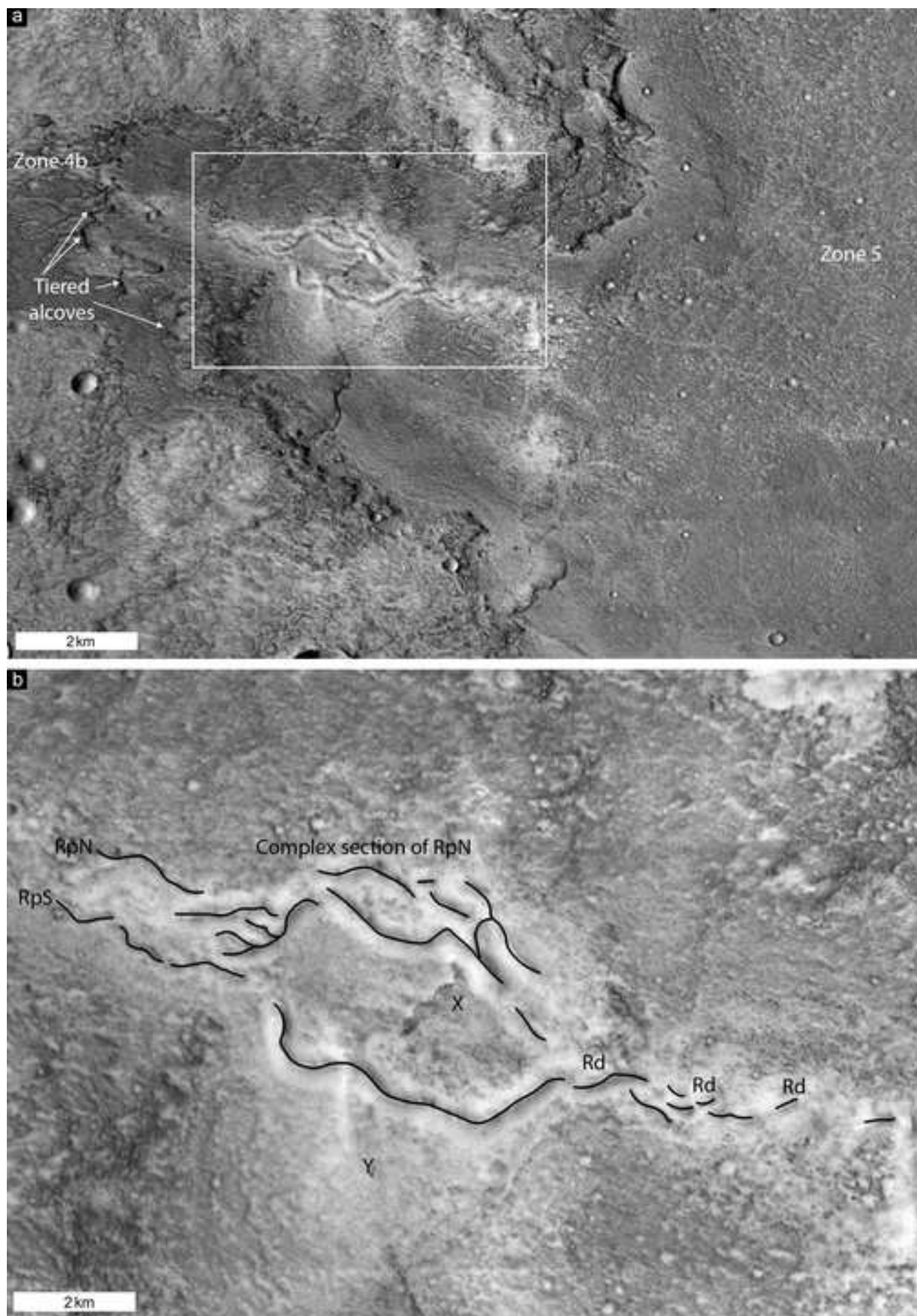


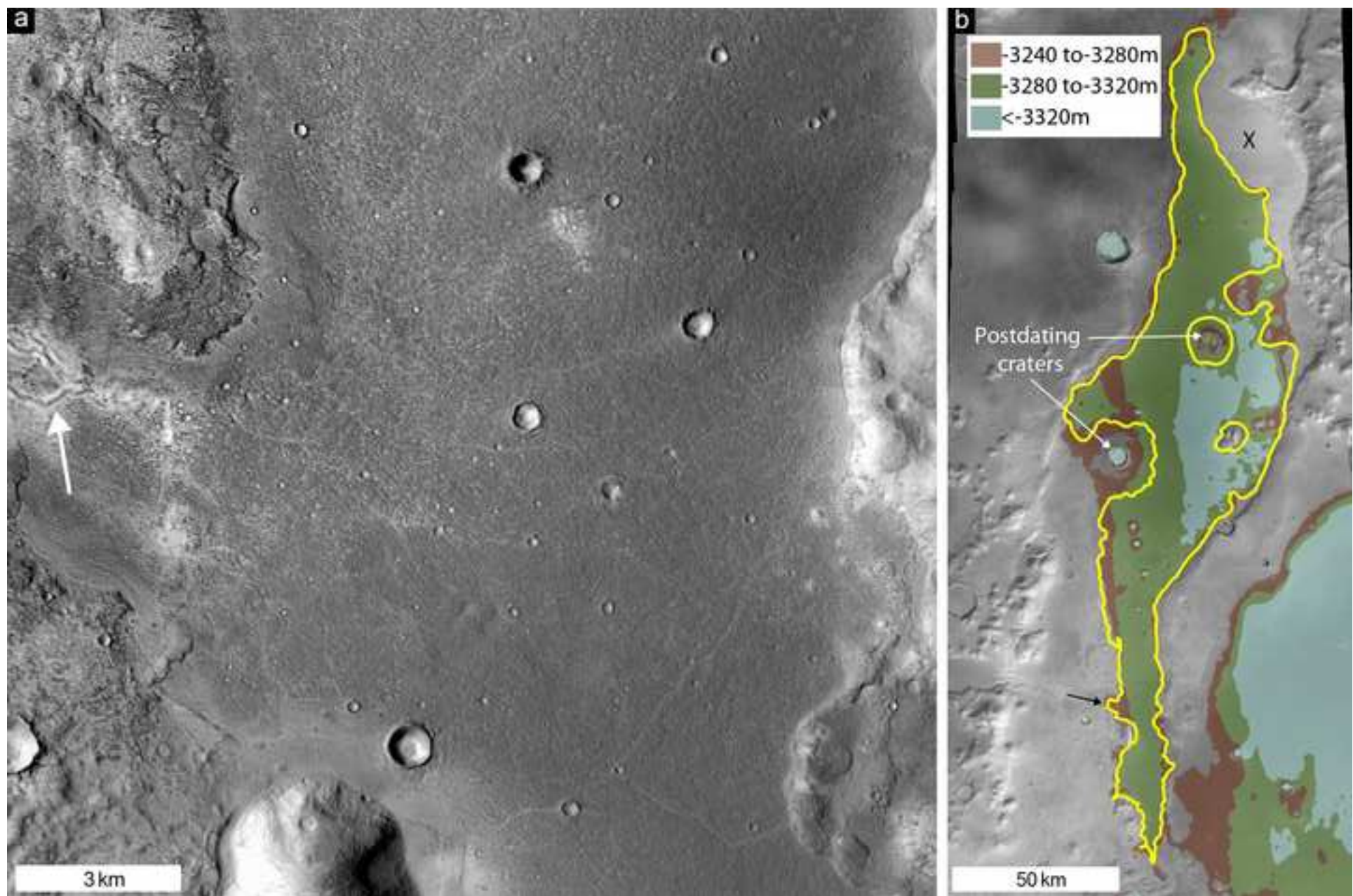
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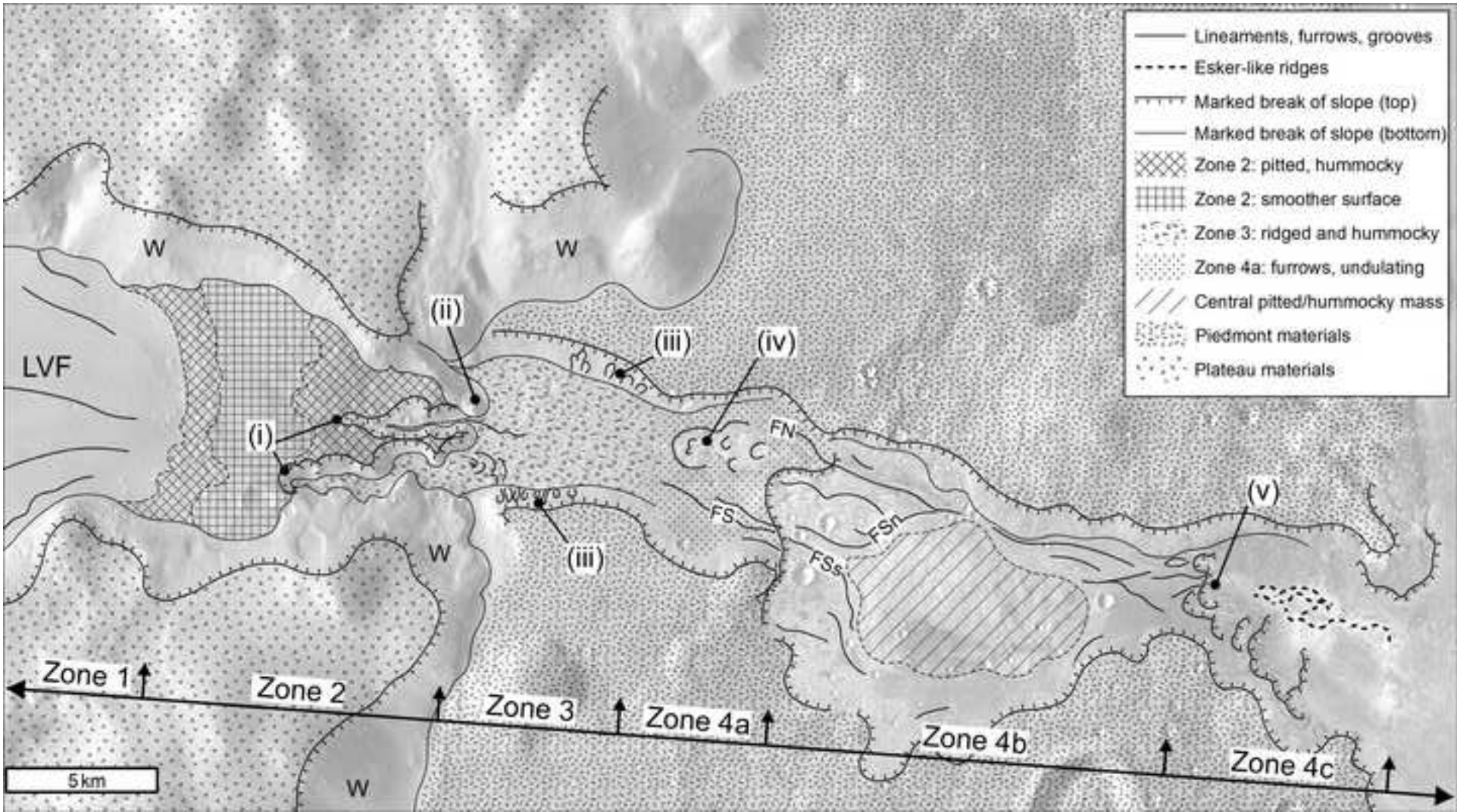
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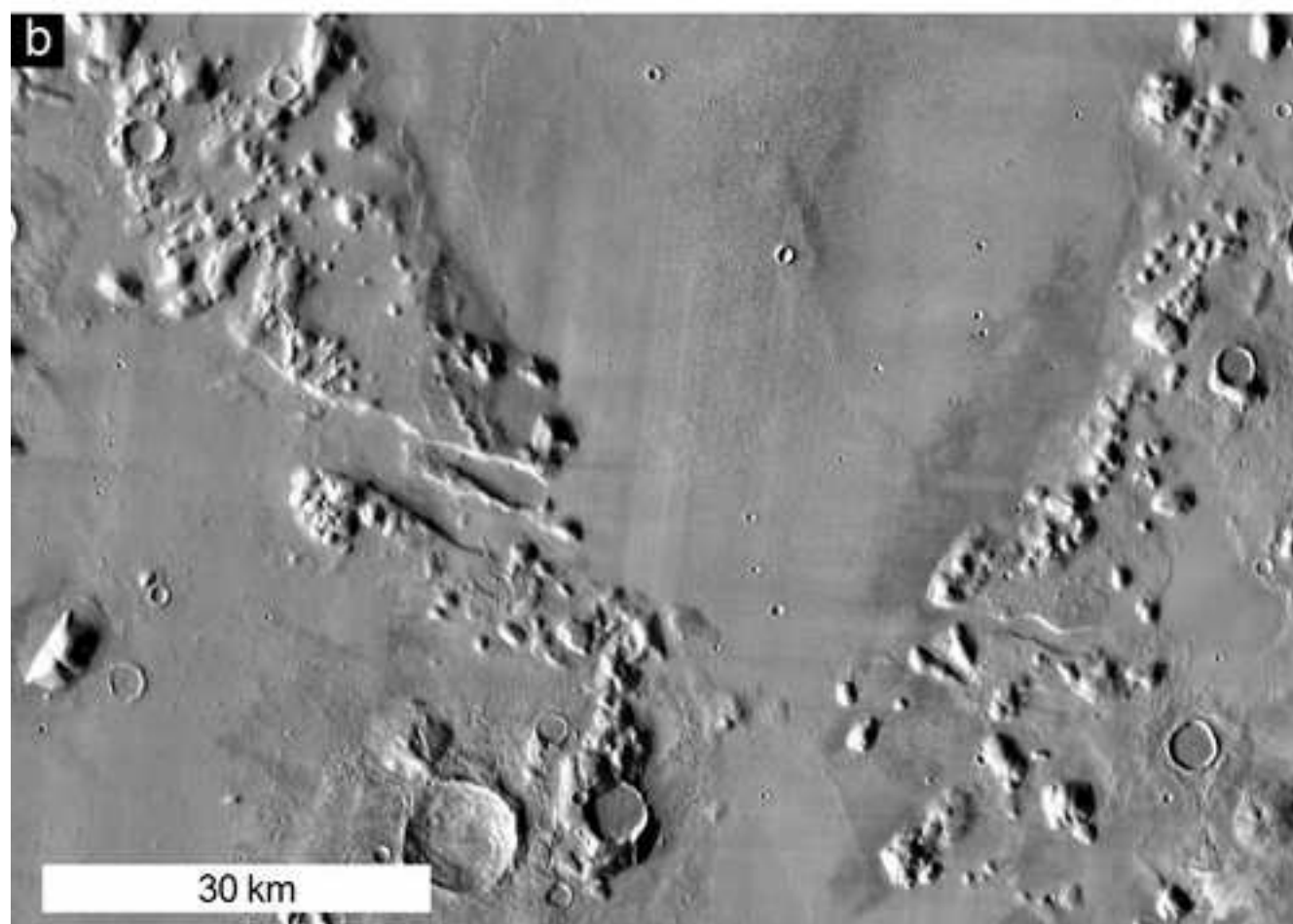
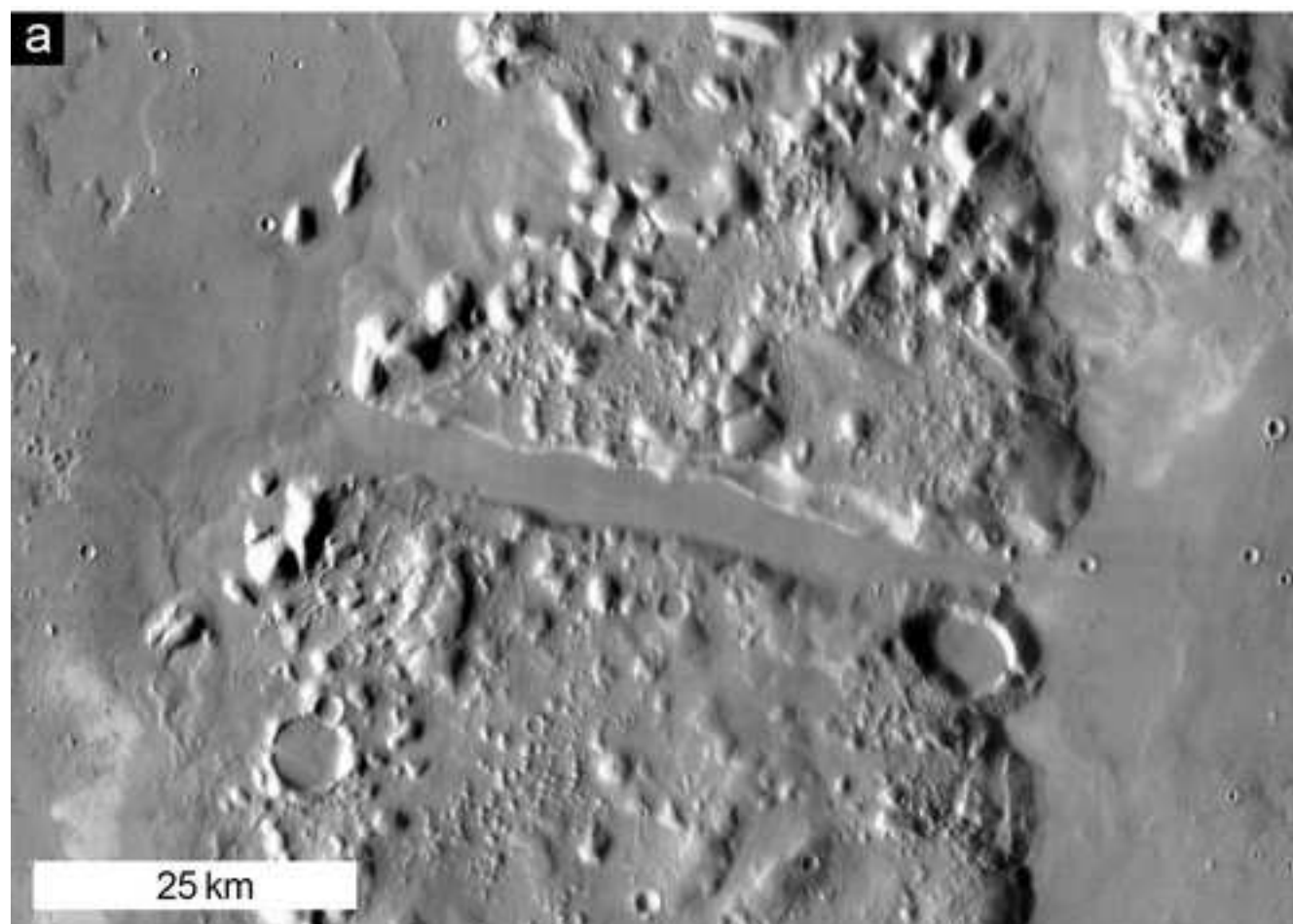
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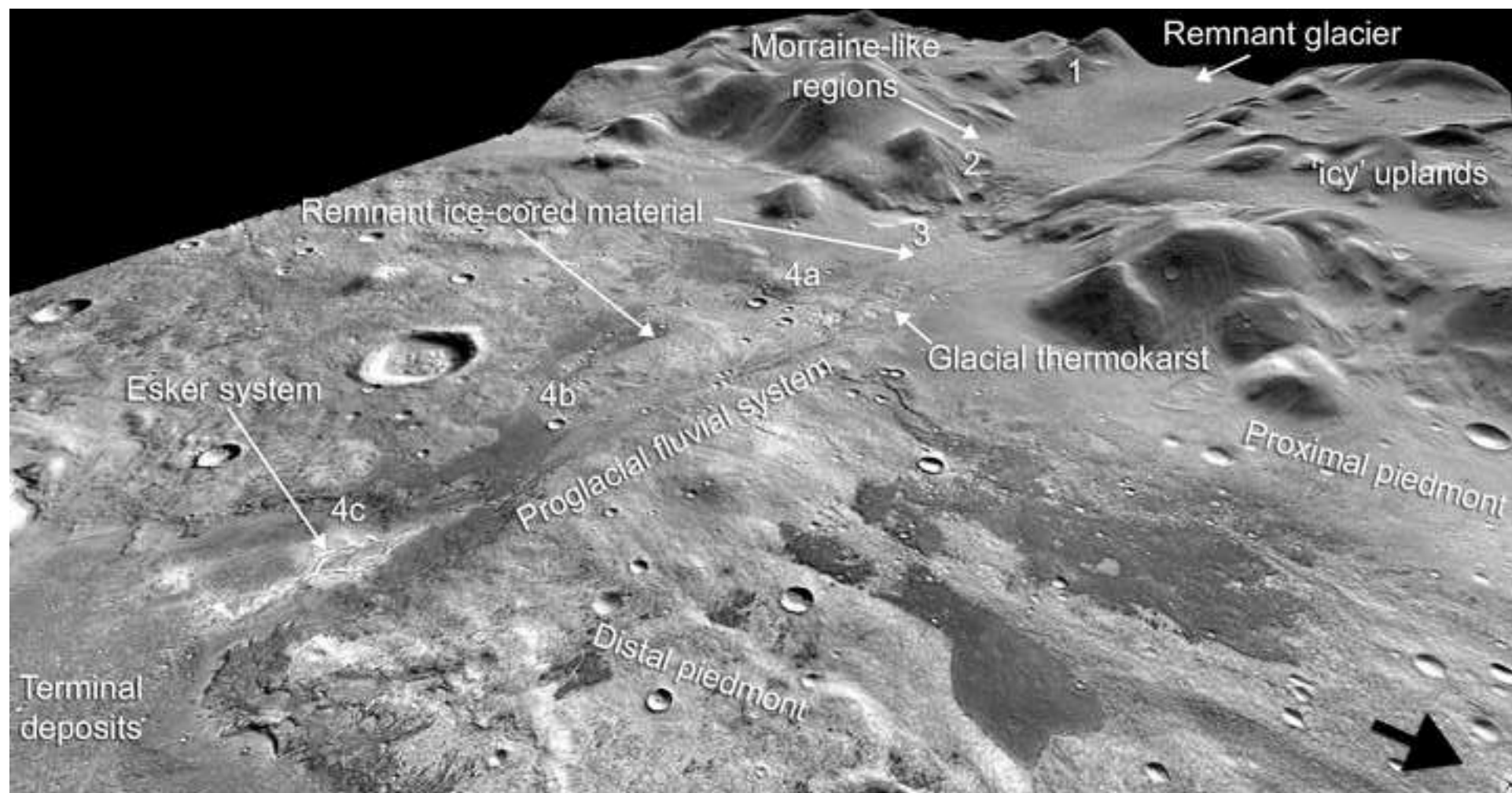
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